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6.—Studies in the Diagenesis of Some Western Australian Sedimentary Rocks

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Textures arising from diagenesis in some Western Australian sedimentary rocks are described, and particular emphasis is placed on determination of the order in which the diagenetic minerals have been developed. Rocks studied in detail include specimens from the Clark Sandstone, Septimus Limestone, Cockatoo Sandstone, Birdrong Formation, Arrowsmith Sandstone, Enokurra Sandstone and Mokadine Formation, and many others are described briefly.

As recommended by Gilbert and Turner (1949) the universal stage has been used to clarify relationships between adjacent authigenic minerals, and to obtain optical and morphological data. It has enabled fairly detailed investigation of some optical properties of K-feldspars in the Clark Sandstone, Mokadine Formation and Arrowsmith Sandstone. The stage is particularly useful for the study of textures in which relatively hard and soft minerals are adjacent, for example quartz and calcite or feldspar and barite. Such mineral pairs are commonly separated by faces of one of the minerals, and these faces, which can be recognized and determined rather easily with the stage, are often otherwise overlooked.

An attempt has been made to summarize the main causes of diagenesis, and to categorize resultant textures according to their origin. Some textures do not conclusively demonstrate the order of formation of the authigenic minerals, but many do, and the latter are almost as useful petrologically in sedimentary rocks as standard or well-known textures are in igneous and metamorphic rocks. The diagenetic textures are divided into (a) enlargement textures, which include simple enlargement textures, indentation textures and enclosure textures, (b) pressure-solution textures, (c) micro-drusy textures, which include simple and composite micro-drusy textures, (d) reorganization textures and, (e) replacement textures.

Reference is made to much of the literature on sedimentary petrology in Western Australia.

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Introduction

The study of sedimentary rocks in thin section has received increasing attention during the last fifteen or twenty years. Contrary to earlier practice, general petrographic texts now tend to allot sedimentary rocks as much space as igneous and metamorphic rocks, and several new texts devoted entirely to the petrology of sedimentary rocks have appeared. This is not to say that sedimentary petrology has reached the position of igneous and metamorphic petrology, in which perhaps most advances in our understanding of petrogenesis come from attempts to duplicate in the laboratory conditions under which the rocks and their constituent minerals form. Thus, although increasing attention is being paid to synthesis of minerals common in sediments, and to their response to various physico-chemical conditions, there is still scope for much petrogenetic inference from microscopic studies of the naturally occurring sedimentary rocks.*

One of the most rewarding avenues for such studies is in the investigation of the minerals and textures arising from diagenesis. A surprisingly complex diagenetic sequence may be unravelled by detailed examination of a rock whose petrogenesis is at first apparently simple.

Use of the Term Diagenesis

Diagenesis has been variously defined. According to Pettijohn (1957, p. 648) the term refers primarily to reactions which take place within a sediment between one mineral and another, or between one or several minerals and interstitial or supernatant fluids. Most authors agree that diagenesis is a low-temperature phenomenon that grades with increase in temperature, into metamorphism. In this paper diagenesis will be divided into early and late diagenesis, though no attempt will be made to define their limits precisely.

Siever's discussion of early diagenesis seems the most satisfactory. Siever (1962, p. 140) considers early diagenesis to include that stage in the history of a sediment during which it is buried only up to a few tens of feet. It is bacterially active, is not greatly compacted, has high porosity and water content, may be moderately permeable, and is subject to the upward passage of waters from compacting sediments below.

Changes attributed to late diagenesis may persist until weathering of the exposed rock: in fact the processes of diagenesis and weathering, though distinguished by many authors, must be gradational.

There is much petrographic evidence for the reality of the distinction between early and late diagenesis, but two examples will suffice here. Calcareous concretions that preserve fishes (see Weeks 1957) are obviously of early origin. On the other hand Topkaya (1950) describes Jurassic conglomerates from Arvel in Switzerland which are free of authigenic silicates despite the fact that they have been partly derived from Triassic formations now exceptionally rich in them. This is evidence of late diagenesis, for the authigenic minerals could have formed only after the erosion of the Triassic rocks.

* It is strange that sedimentary petrology should have lagged behind igneous and metamorphic petrology. The first rock to be examined in thin section with a microscope and polarized light was a sediment (Sorby 1851), and Sorby later published widely on sedimentary petrology, maintaining an interest for many years in such contemporary sedimentological problems as dolomitization and the formation of aragonite (Sorby 1904).

One of the main problems has been to decide when diagenesis becomes metamorphism. It has been suggested by Packham and Crook (1960, p. 404) that retention of the original clastic fabric of the rock would indicate that it had been subjected only to diagenesis, whereas extensive modification, involving substitution of hornfelsic or schistose fabrics, would demonstrate metamorphism. There is no difficulty in distinguishing diagenesis from metamorphism in most of the rocks discussed below. Some of them, as indicated later, have been intruded by dolerite. Apart from that, none has been metamorphosed in the sense of having been heated by intrusions, strongly folded or generally sheared; nor are any adjacent to later igneous bodies that might have metasomatized them. Further details about their age, location and tectonic setting can be obtained from McWhae *et al.* (1958).

Aims of the Investigation

The purpose of this paper is to show how texture may demonstrate the order in which diagenetic processes have taken place, and how it can reveal some of the causes of those processes. This has been done before, notably by Gilbert (1949) and Gilbert and Turner (1949), and comprehensive reviews of the literature on diagenesis have been given by Pettijohn (1957) and Carozzi (1960). Although the techniques used here are similar to those of Gilbert and Turner, there are several reasons, stated immediately below, why presentation of the results of this study is warranted. Firstly, reference is made to all known papers dealing with diagenesis in Western Australian rocks. Secondly petrographic descriptions of some of the sedimentary rocks treated here have not previously appeared in the rather sparse Western Australian literature. Thirdly, some of the textures are different from those described elsewhere. Finally, an attempt is made to group the various diagenetic textures with similar origins into categories, so that their petrologic interpretation is thereby facilitated.

The Universal Stage in Sedimentary Petrology

General Advantages

Gilbert and Turner point out that the universal stage is particularly valuable in the microscopic investigation of sedimentary rocks, and state that it is essential in a modern sedimentary laboratory. They use the stage to determine the optical properties of the minerals, and the crystalline faces developed, and to observe the three-dimensional interrelationships of the minerals. Faces are identified tentatively by noting their relationship to optical directions within the crystal, and their identity is confirmed by measuring angles between the faces and comparing them with published data, notably those in Dana (1899). Gilbert and Turner note, as was found in the present investigation, that appropriate tilt on the stage reveals crystalline faces quite unsuspected following routine microscopic examination of authigenic minerals. They recommend the Federow procedure in which all measured directions and planes are plotted on a stereographic or equal-area projection net. Details of procedures are clearly set forth by them, and will not be repeated here.

Detection of Crystal Faces

Where the boundary between two authigenic minerals is planar, the plane is commonly a face of one of the minerals. The fact that a boundary is planar can only be seen if the plane is essentially parallel to the microscope axis: if it is oblique the boundary does not appear sharp and is represented by a narrow band in which the two minerals, perhaps with very different optical properties, overlap. Moreover, unless both minerals have a similar resistance to abrasion, irregularities due to grinding can cause an inclined planar boundary to appear uneven. This effect, which is common with such mineral pairs as quartz and barite, quartz and calcite, and quartz and dolomite, is illustrated in Fig. 1, and Plate I, Figs. 1 and 2. The only reliable way to establish the planar nature of boundaries between such minerals and hence to discover crystal faces is by tilting until the plane is parallel to the microscope axis.

Examination of Replacement Textures

Where carbonate cement completely encloses quartz grains it commonly penetrates their margins and partly replaces the quartz. On the other hand, the margins of well-rounded grains that have undergone no replacement may show apparent irregularities for reasons indicated in the previous paragraph and these irregularities may suggest marginal replacement. Appropriate tilt, revealing smooth, rounded margins devoid of re-entrants, prevents misinterpretation.

Diagenetic Minerals and Textures

Textures resulting from diagenesis are considered below, the constituent minerals are described, and the order of their formation is discussed. Certain diagenetic textures are commonly associated with particular lithologies, and some textures are virtually restricted to one lithologic type. For this reason, the nature of the textures is illustrated by describing the mineral relationships commonly encountered in some of the main types of sedimentary rock. Discussion of the ultimate cause of diagenetic growth and solution of minerals is however left to a later section of this paper.

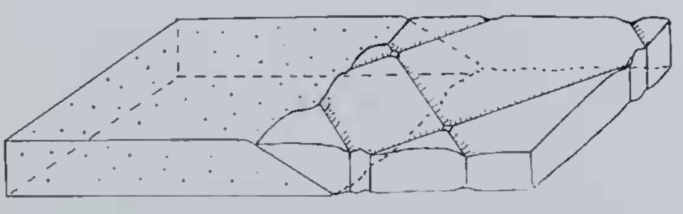


Fig. 1.—Diagrammatic sketch of unmounted thin section of quartz-barite rock. Quartz (stippled) and barite (cleaved) meet in a plane. However, grinding has lowered the level of the softer barite with respect to quartz, and has plucked out cleavages, forming rough valleys. The quartz-barite boundary, viewed from above, appears highly uneven. The planar nature of the contact will be evident only if the section is tilted so that the plane is parallel, or almost parallel, to the microscope axis.

Feldspathic and Arkosic Sandstones

Clark Sandstone.—The Clark Sandstone (Traves 1955) of the Carlton Basin, in north-eastern Western Australia, is a highly fossiliferous, reddish or green, glauconitic, Cambrian sandstone. It is faulted but not strongly folded, and it is not intruded by later igneous rocks. The preservation of abundant brachiopods and trilobites, and much unaltered glauconite, shows that the remarkable outgrowths on quartz and feldspar observed in this study must be ascribed to diagenesis rather than metamorphism.

The specimen examined (No. 48629*, from the Clark Jump Up, on the Carlton-Legune track, about 35 miles east of Wyndham) is a weakly lithified, porous, well-sorted, medium-grained sandstone with the following composition (by volume): quartz 63%, glauconite 16%, K-feldspar 9%, rock fragments (fine-grained quartzite, mica schist, chert) 6%, muscovite $< 1\%$, and a red, mainly haematitic, matrix 6%. Most quartz and feldspar grains, before enlargement, were somewhat rounded, and some were well rounded.

Almost every quartz and feldspar grain has been enlarged, and where there has been space quartz has developed prisms and rhombohedra, and feldspar has developed prisms and basal and side pinacoids. Glauconite, generally regarded as a product of early marine diagenesis, is present as ovoid pellets. The authigenic quartz and feldspar formed after the glauconite, as shown by the way in which they are moulded on glauconite grains upon which they impinged during growth. Many quartz and feldspar grains are therefore partly bounded by crystal faces, and partly by rounded concave surfaces where they abut the glauconite. Some of the glauconite grains have yielded by gliding on micro-faults within the pellets, apparently because of squeezing and fracturing by the force of the growing quartz and feldspar (Plate I, figs. 3 and 4). This indicates that the authigenic quartz and feldspar formed after at least moderate compaction of the sandstone, for in an unconsolidated sand the glauconite pellets would have been moved aside rather than broken.

The cores of the K-feldspar grains are adularia ($2V_\alpha$ from $57\frac{1}{2}°$ to $68°$) or microcline (except for one unusual grain with a composite core of adularia and microcline). Cleavage can commonly be seen passing without break or change of direction, from both types of core to the authigenic rim, but any twinning in the core is sharply truncated at its boundary with the rim (Plate I, Figs. 5 and 6). The 001, 010 and 110 cleavages and faces were identified by measuring the angles between their normals and X, Y and Z, and comparing them with Nikitin's data (see Gilbert and Turner 1949, Table IV). These measurements, which were made with the universal stage, also served to confirm the identity of the core, for they enable distinction between untwinned microcline and adularia.

The authigenic rims have remarkable optical properties, but due to their narrowness, numerous inclusions and patchy extinction, precise

* Numbers refer to the General Collection of the Department of Geology, University of Western Australia.

C3571—(2)

measurements are possible only on selected grains. Patchy extinction is not accompanied by obvious bending or distortion of cleavage. The angles between the principal optical directions and the normal to 010 and 001 correspond approximately with those quoted by Nikitin for orthoclase. This correspondence can clearly not be everywhere precise in rims with patchy extinction, for the patchiness is caused by variations of up to 6° in the principal optical directions. Thus Z commonly departs by several degrees from the normal to 010 and the rims are not everywhere monoclinic. Moreover, there is generally a variation of 2V within the one rim. The maximum variation observed within one rim is from 11° to 45.5°, and the overall range is from 11° to 64° (see Table I). This range of 2V in K-feldspar would thus include the minerals, as sometimes defined, adularia ($2V\alpha$ = 50° – 70°), orthoclase ($2V\alpha$ = 25° – 50°), and sanidine ($2V\alpha$ = 0° – 25°, optic plane ⊥ 010) [see Chaisson (1950) for the optical classification used here]. No grains whose optical plane is parallel to 010 (i.e. "high" sanidine) were observed.

All of the authigenic feldspar described above came from one hand specimen 2 in. x 2 in. x 1 in., and the considerable range in 2V between different grains, and within individual grains, is puzzling. Variations in the optical properties of potassic feldspars are said to depend upon (1) compositional differences, (2) submicroscopic intergrowths due to unmixing, and (3) degree of Si/Al order (see Hewlett 1959). Neither the composition of intrastratal solutions at any given time, which influences the composition of the feldspar precipitated, nor the cooling history of the feldspar, which affects the degree of unmixing, and of Si/Al order, can have varied much. This suggests the influence of other, highly localized factors. Differences between the clastic cores, which are seed crystals, could cause precipitation of slightly different material on each. Perhaps slight compositional differences within one core could cause or even accentuate, differences in material precipitated in different parts of the one rim. The shape and volume of the intergranular pores may have affected the rate of passage of solutions in a significant way. This, however, is speculation: no reason can confidently be advanced, at present, for the variations in 2V between different authigenic rims, and within individual rims. No explanation is offered, either, for the remarkably low values of some readings (see Table I).

Mokadine Formation.—The Mokadine Formation, of probable Proterozoic age, crops out on the eastern margin of the Perth Basin near Moora. It is a sequence of arkose, feldspathic sandstone, siltstone, and claystone that has been intruded by dolerite dykes. Authigenic growths on detrital quartz grains are mentioned by Logan and Chase (1961) in their brief description of the petrography of the formation. The specimen described in this investigation (No. 36909, collected from just above the type section of the underlying Dalaroo Siltstone) is a brown, strongly lithified, moderately well-sorted, fine-to-medium-grained arkose and it does not appear to have been altered by dolerite intrusions, the nearest exposed dyke being ¼-mile distant. It has the following composition (by volume): quartz 65%, K-feldspar 25%, opaque iron-ore grains, largely changed to haematite, 8%, and interstitial haematite and limonite, 2%. The quartz includes a few fine-grained quartzite grains. The K-feldspar includes microcline and adularia ($2V\alpha$ down to 58°), and ranges from clear fresh grains to grains that have been completely converted to grey and yellow-brown clay minerals. The interstitial iron oxide is present mainly as a film that outlines the clastic grains, which range from sub-angular to very well rounded.

TABLE I

Variations in the optic axial angles of authigenic feldspar in the Clark Sandstone. Values greater than 25° are reproducible to within 1°, and values less than 25° are reproducible to within 2°.

Clastic Core	$2V\alpha$ of authigenic rim
Microcline ...	17°
Microcline ($2V\alpha$ = 80°) ...	47°, 37½°
Adularia ($2V\alpha$ = 68°)	24°, 22½°
Microcline ($2V\alpha$ = 82½°) . .	64°
Adularia ($2V\alpha$ = 61°)	45½°, 43°, 24°, 11°
Microcline	41°, 18°
Microcline intergrown with adularia ($2V\alpha$ = 57½°)	43½°, 41½°
Microcline	25½°

Almost all quartz and many K-feldspar grains have been authigenically enlarged. Secondary quartz is abundant, forming about 5% of the rock, and secondary K-feldspar makes up perhaps 0.5%. The following points are significant:

1. A few clastic quartz grains penetrate each other to give minor and poorly developed microstylolitic intergrowths. Quartz grains commonly penetrate feldspar grains by as much as 0.025 mm without development of microstylolitic boundaries, and the feldspar has apparently undergone almost all the solution in such pairs (see Plate II, Fig. 1). Much less frequently a fairly sharp-cornered feldspar grain slightly penetrates the flattish surface of a quartz grain, showing that initial shape and perhaps crystallographic orientation influence the final relationship of the two minerals. Feldspar grains are also penetrated by each other. No clastic grains have been penetrated by authigenic outgrowths, so compaction preceded authigenesis, which was a late diagenetic feature. The volume of secondary material is far greater than that available from local solution by clastic grains at their point of contact, and most of it must have come from elsewhere.

2. Boundaries between quartz outgrowths are either irregular, or roughly planar, but the planes generally do not seem to correspond to common rational faces of either grain. Outgrowths of quartz and feldspar meet in boundaries that occasionally correspond to the 001 face of the feldspar, but are more often approximate planes that are not rational faces of either grain. These relationships may mean that times

of growth of the minerals overlapped (see Gilbert and Turner 1949, p. 13).

3. Each feldspar growth is kaolinized to about the same degree as its clastic core. This shows that most kaolinization occurred after formation of the authigenic rims, so that the clastic feldspars were practically unaltered when deposited. Kaolinization has apparently lowered the optic axial angle of some grains, for some readings on clastic microcline cores are as low as 72°, about 10° lower than those normally quoted for the mineral. As the rims are also altered, their range in $2V_\alpha$ from 68° to 77° gives no reliable indication of their nature.

4. Many authigenic rims on microcline grains have uneven extinction, apparently a vague continuation of the crosshatching (albite and pericline twins) of the core. It has been suggested that crosshatching indicates inversion from an earlier, high-temperature monoclinic form (Laves 1950), and the fact that authigenic (low-temperature) microcline described by Baskin (1956) does not show crosshatching, seems to accord with this view. Baskin's grains show no clastic cores and have a type of fourling twin, and their unusual morphology may well result from cold formation. On the other hand the poorly developed extension of crosshatching from clastic cores to rims in microcline of the Mokadine Formation does not necessarily mean that they grew at high temperature. These rims seem to have their structure determined by that of the core rather than by temperature of formation, for they are crystallographically continuous with the adjacent part of the core. The same effect is observed with strained quartz grains, whose rims show wavy extinction continuous with that of the adjacent quartz, and with plagioclase, the twinning of which extends from the core into the rim (see Plate II, Fig 4). The vague extension of crosshatching noticed here therefore probably has no significance in terms of temperature. However, if the type of crosshatching just described is not an indication of high temperature formation, it is strange that it is not more common.

It is not known whether authigenic quartz and feldspar replaced an earlier mineral cement, or whether they grew in voids between the grains. Logan and Chase (1961) found specks of undigested calcite in the authigenic quartz cement of arkoses in the Mokadine Formation examined by them, but no calcite could be detected in this rock. The only incorporated materials are the iron oxide coatings outlining the clastic grains.

Arrowsmith Sandstone.—The Arrowsmith Sandstone is an unfossiliferous, lower Palaeozoic or Proterozoic formation exposed in the north-eastern part of the Perth Basin near Arrino. All specimens so far examined by the author are well-sorted, medium-to coarse-grained arkoses, and many of them contain almost as much lithic material as feldspar, and therefore they nearly grade into lithic sandstones. The lithic fragments include sericite-quartz schist, composite quartz-feldspar grains, and myrmekite, but the most abundant are volcanic fragments. A previous account of the petrography (Glover 1960*b*) drew attention to authigenic quartz and chlorite, but re-examination with the universal stage has revealed a complex diagenetic sequence.

Specimen Pf2* is an arkose with the following composition (by volume): quartz 40%, feldspar 27%, volcanic fragments 13%, other lithic fragments 13%, authigenic cement 6%, other minerals 1%. The grains are angular to well rounded, and there is slight pressure-solution on some boundaries of adjacent grains, causing penetration of one by the other, generally without discernible micro-stylolitic intergrowth.

Almost all clastic grains are covered by a thin brown film that generally appears as a very faint stain on the grain surface. This film is only 0.002 mm thick in cross section, and is composed of one, or in places, two or three layers or flakes of a brown mineral with fairly high relief. The mineral is pleochroic from red-brown to yellow, with absorption greatest when the flakes are elongated parallel to the polarizer, but it is too thin for other optical properties to be measured or even clearly observed. The mineral seems most closely allied to brown mica, and it may represent the transformation product of a thin argillaceous coating on the sand grains. It has disappeared locally where pressure-solution has caused slight interpenetration of clastic grains.

Other authigenic minerals beside the brown, mica-like mineral, are quartz, pale green almost isotropic chlorite, and feldspar. The feldspar is untwinned, with relief less than quartz, and usually has patchy extinction. $2V_\alpha$ ranges between 43° and 55° with an average of 50° (10 grains) and typical measurements of the angles between X, Y and Z and the pole of the main cleavage are as follows (the corresponding angles between 001 and X, Y and Z for orthoclase, as given by Nikitin, are quoted in brackets):

X $\wedge$ $\perp$ cleavage = 81.5° (85°).
Y $\wedge$ $\perp$ cleavage = 9° (5°).
Z $\wedge$ $\perp$ cleavage = 88.5° (90°).

The feldspar is close to the orthoclase-adularia boundary, according to the classification of Chaisson (1950). It is uncertain whether the differences between the optical measurements given above and those of orthoclase taken from Nikitin are due to experimental errors arising from the smallness of the grains and their uneven extinction, or whether there is a slight departure from monoclinic optics.

The authigenic minerals listed occupy the spaces between clastic grains, and examples of the resultant diagenetic textures are illustrated below. Plate II, Fig. 2, shows one of the simplest arrangements. The minerals lining the void are a brown, mica-like mineral on the surfaces of the clastic grains, chlorite which has a serrated fringe toward the centre of the void as though it had grown inward, and feldspar in the centre. This texture would not arise from replacement of an earlier cement, but is consistent with precipitation in an empty pore space. As with similar drusy textures, which form by precipitation from fluids in cavities, the first-formed mineral must be the outermost, and the last-

* Specimen number assigned by West Australian Petroleum Pty., Ltd.

formed the innermost. In the texture illustrated in Plate II, Fig. 2, the brown, mica-like mineral formed first, the chlorite second, and the feldspar last. Some of the intergranular micro-drusy textures are more complex, and there have been minor variations in the order of formation of the minerals: nevertheless, the basic pattern is the same, and the brown mica-like mineral is always the first, with chlorite either second or third.

In some places there is a similar texture in which quartz is substituted for feldspar. In the texture illustrated in Plate II, Fig. 3, quartz forms both before chlorite and after it. The most complex texture is shown in Fig. 6b, where quartz forms before chlorite and both quartz and feldspar form after it. The last-formed quartz has perfectly developed prism faces and, as the mineral was growing into a void, it must have formed before the feldspar. The order of development in this texture was therefore (1) brown mica-like mineral, (2) quartz, (3) chlorite, (4) quartz, (5) feldspar.

Elsewhere in the Arrowsmith Sandstone similar textures, but with development of slightly different minerals, are encountered. The composition of the chlorite changes, and its colour and birefringence vary perceptibly. The texture shown in Plate II, Fig. 4, is notable, for it shows how albite twinning of plagioclase [determined by the curves set forth by Turner (1947) as An_5] persists into outgrowths. The original clastic grain is outlined by the brown micaceous mineral.

Any reasonable conjecture regarding the origin of the diagenetic textures in the Arrowsmith Sandstone must seek to account for the presence together of up to four minerals in the spaces between the clastic grains, and should explain their sequence and arrangement. The following hypothesis appears to satisfy these conditions. A well-sorted sand, with most of the mud winnowed out to deeper water, is assumed to have compacted to a porous sandstone. Slight pressure-solution causing minor penetration of clastic grains accompanied this compaction, or was brought about by later earth movement. The thin film adhering to the grains was dissolved at their points of contact: it is not clear when this mineral, probably originally argillaceous, was changed to the brown mica-like mineral that now coats the grains. In any case, in one form or another, it must have preceded the diagenetic chlorite, quartz and feldspar, which were precipitated from percolating waters. These waters, which varied in composition both in time and place, were magnesian, potassic, sodic and siliceous, and all the solutes would have been available within the formation itself, as it contains volcanic fragments, microcline, orthoclase, sodic plagioclase and quartz. Precipitation could have taken place at any time after compaction of the sediment in the late Proterozoic or early Palaeozoic. It is conceivable that equilibrium between the fluid and surrounding rock was established several times, only to be disturbed by flowage of the interstitial fluids caused by earth movements. If so, processes leading to attainment of the final texture might have persisted even for millions of years. It might indeed be speculated whether future techniques will allow absolute dating of authigenic minerals in such textures. The dates obtained could well apply to tectonic events, such as folding or uplift, that brought about the disequilibrium mentioned above.

Enokurra Sandstone.—The Enokurra Sandstone is an unfossiliferous Proterozoic or lower Palaeozoic formation exposed in the north-eastern part of the Perth Basin, near Yandanooka. It consists of fine- to very coarse-grained sandstone which is locally conglomeratic, and shows well-developed cross-bedding. It has been described previously by McWhae *et al.* (1958), Glover (1960*b*) and Bastian (1961).

The specimen investigated here (No. 38725, from 3 miles northeast of Yandanooka) is a coarse-grained arkose with the following approximate composition (by volume): quartz 67%, feldspar (microcline and oligoclase) 15%, lithic fragments (mainly gneiss) 14%, other minerals (muscovite, garnet, opaque grains, clay-sized material, (?) limonite, haematite) 4%. The clastic grains are outlined by a patchy coating of haematite. Cementation has been effected partly by muscovite and patches of sericitic to clay-sized material, and partly by secondary quartz. The secondary quartz fills many of the spaces between the clastic grains, and is usually in optical continuity with one or more of the visible quartz grains. In places, outgrowths from several grains meet in more or less planar surfaces, but none have so far been identified as rational crystal faces. Also present in these areas of authigenic quartz is a fringe of colourless to very pale-green sericite ($2V\alpha$ 38°) adhering to the brown, iron-stained clastic grain surfaces. The fringe, which is made up of minute flakes at right angles to the grain surfaces, is commonly difficult to observe, even under crossed nicols, for the interference colours of the minute flakes are hard to distinguish from those of the surrounding authigenic quartz (see Plate II, Fig 5).

The texture here is basically the same as that observed in the Arrowsmith Sandstone, and could have arisen only by precipitation in voids between grains. A few of the clastic grains penetrate each other, and some of the contacts are micro-stylolitic. Even if quartz dissolved at the points of contact of these grains were deposited nearby in voids, in the manner envisaged by Waldschmidt (1941), there would have been insufficient available from that source to account for all the secondary quartz in the rock. Moreover, some other source for the sericite would be necessary. Precipitation from potassic and siliceous solutions percolating through the porous rocks was the probable cause of this diagenetic texture.

Quartz Sandstones

One of the most common types of diagenesis is the enlargement of quartz grains, especially in well-sorted quartz sandstones such as some from the Birdrong Formation, Cockatoo Sandstone and Wogatti Sandstone. Diagenetic enlargement of quartz is also common in feldspathic sandstones from the Clark Sandstone, Mokadine Formation, Arrowsmith Sandstone, and Enokurra Sandstone, discussed earlier in this paper. Petrologic descriptions of sandstones in the Billeranga Beds, Wenmillia Formation and Moora Group that contain abundant secondary quartz have been given by Arriens and Lalor (1959), Ranford and Shaw (1960) and Logan and Chase (1961) respectively. More petrographic work will undoubtedly confirm the widespread nature of authigenic quartz in unmetamorphosed Western Australian sandstones: for example recent petrographic descriptions of sediments from the Canning Basin by Johnson and Dallwitz (*in* Veevers and Wells 1961) mention it in the Noonkanbah Formation, Grant Formation, Godfrey Beds and Kidson Beds. Solution of quartz is also an important process in sedimentary rocks, but in sandstones it is most usually found where there is an abundant calcareous, ferruginous or argillaceous matrix or cement. The dissolved quartz is almost always replaced by the cementing material. However, solution processes and their resultant textures are considered elsewhere (p. 44 *et seq.*).

The sandstone illustrated in Fig. 2 shows the type of texture that results from enlargement of quartz grains in a well-sorted quartz sandstone, where the growth does not fill the pore spaces between grains. This sandstone is from the Cretaceous Birdrong Formation in the western part of the Carnarvon Basin. The original rounded shapes of many grains have been outlined, apparently by a coating of the same argillaceous material that now partly occupies the pores. Some of the grains are in direct contact with each other, but there seems to have been no pressure-solution, and the secondary quartz must come from another source. Boundaries between secondary quartz outgrowths in adjacent grains are planar, and all measured planes are prisms or rhombohedra of one or the other of the grains.

Relationships between the grains are worth studying, for the light they may throw on the process of crystallization in a porous sandstone. For example, in Fig. 2 the contacts between grain A and grains D and B are a rhombohedron and a prism respectively of grain A, indicating that growth on these faces had been completed before the quartz of grains D and B grew out to meet them. Similarly, grain D finished growing before grain C along their common boundary, and grain C in turn finished before grain B. Thus the sequence, beginning with the first to complete its growth, is A, D, C, and B. The position in this sequence of grain E is unknown, except that it finished crystallization before adjacent parts of grain C. The sequence inferred above does not imply that individual grains accomplished all their growth one after the other, during separate and distinct intervals of time. Their growth almost certainly overlapped, as the rock was sufficiently porous to allow fluids to percolate widely.

An example of quartz sandstone in which silicification has almost completely filled spaces between the grains is specimen 48624 from the Devonian Cockatoo Sandstone of the Bonaparte Gulf Basin. The formation is made up mainly of cross-bedded sandstones, and specimen 48624 comes from Cockatoo Springs, about 65 miles

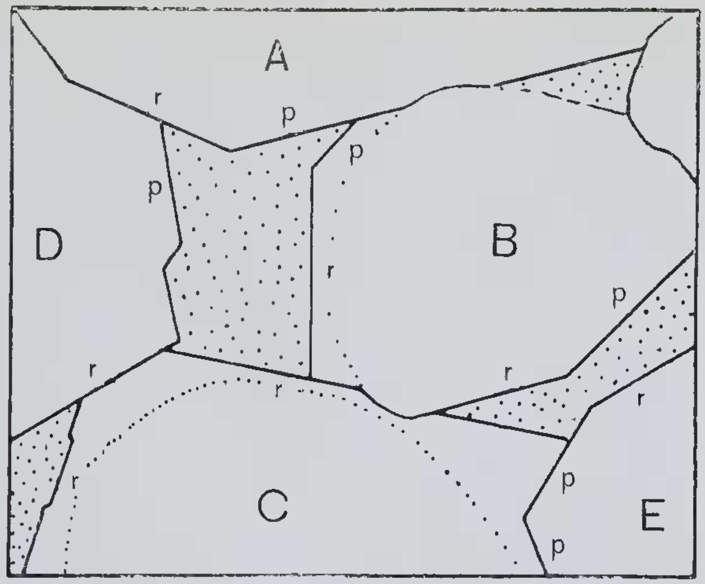


Fig. 2.—Authigenic enlargement of quartz in sandstone from the Birdrong Formation (Rough Range Bore No. 1, core 7, 3,633-3,636 ft). The stippled areas are partly filled with clay-sized material. Quartz crystal faces are indicated by p (prism) and r (rhombohedron). The sketch is slightly idealized to show faces clearly. Width of field 0.6 mm.

east-south-east of Wyndham. It is made up of rounded quartz grains with a haematite coating (85%), authigenic cement (11%) and other minerals including microcline and adularia with authigenic outgrowths, quartzite fragments, green tourmaline, muscovite, and kaolinized composite grains. Numerous planar boundaries can be identified as common faces of quartz, but other boundaries are not strictly planar; of the latter some almost correspond to rational faces, but many cannot be identified. Some apparently irregular or curved faces appear to be made up of many very small planes. This texture is apparently that which would have developed in sandstones of the Birdrong Formation, had silification proceeded to the point where most outgrowths interfered with each other. The tendency for the development of some rational crystallographic faces, despite mutual interference, has also been observed in sandstones studied elsewhere (Basumallick 1962).

Most authigenic outgrowths on the rare K-feldspars have developed faces, among which 001, 010, 110 and 130 have been identified. Some of the feldspar outgrowths abut the clastic cores of the quartz grains, and have been moulded by them. This is the same as the texture illustrated in Fig. 5e, and is proof that the authigenic feldspar preceded the authigenic quartz in this rock.

Calcareous and Dolomitic Rocks

Many clastic limestones are mineralogically and texturally simple. On the other hand, numerous calcareous and dolomitic rocks, ranging from calcareous or dolomitic quartz sandstones to limestone, contain complex suites of diagenetic minerals, of which the most usual are pyrite, glauconite, quartz, opal, chalcedony, albite, orthoclase, microcline, calcite and dolomite. The order of formation of these minerals may often be established by their textural relationships. Some textures, and thus some corresponding sequences of formation, appear to be more usual than others. The reason for the common presence of authigenic siliceous minerals in carbonate rocks has been debated for a long time, but satisfactory explanations are only now being advanced. Textures from Western Australian calcareous and dolomitic rocks will now be considered.

Clastic Limestones.—Our understanding of the genesis of clastic limestones has increased greatly in the last ten years, particularly with the research into the formation of Recent limestones on the Bahama Banks by Illing (1954), and later workers. The classification of Folk (1959) incorporates much of the new information on the significance of textures in these rocks. The only published petrologic work on Western Australian calcarenites (Glover 1955) deals in introductory fashion with Devonian rocks from the Lennard Shelf, Canning Basin, but was done before the results of Illing's work were known. Devonian clastic limestones of the Canning Basin, because of their abundance, strong outcrop, and lack of metamorphism, would be especially suitable for the type of detailed quantitative petrographic investigation carried out by Stauffer (1962). Moreover, they are associated with well-exposed reefs and in places provide excellent examples of dolomitization for study.

Two clastic limestones are illustrated here to show textures arising from partial cementation of a calcarenite from the Pleistocene Coastal Limestone at Shark Bay (Plate III, Fig. 1) and complete cementation of an oolite from the Devonian of the Lennard Shelf (Plate III, Fig. 2). Both textures are formed by precipitation of calcite from solutions in voids, and can readily be distinguished from textures due to recrystallization of a fine matrix (Fig. 6e).

Dolomitic Quartz Sandstone.—Specimen 43799 is from the Palaeozoic sequence in the Bonaparte Gulf Basin and is a fine-grained, silicified, dolomitic quartz sandstone containing quartz (88%), dolomite (10%), pyrite (2%), and traces of other minerals (zircon, tourmaline, black iron-ore, and microcline). The

Fig. 3.—Indentation texture in dolomitic quartz sandstone (No. 43799). Dolomite is partly surrounded by authigenic quartz. The clastic core of the quartz grain is clearly shown. This texture does not reveal whether the dolomite or quartz grew first. Crossed nicols. Width of field 0.25 mm.

Fig. 4.—Indentation texture in dolomitic quartz sandstone (No. 43799) showing special case where dolomite is partly surrounded by authigenic quartz, but has been moulded onto the clastic core. Dolomite grew before the authigenic quartz. Crossed nicols. Width of field 0.25 mm.

clastic quartz grains are fairly well sorted and generally well rounded, and the sandstone is undoubtedly a mature, multicycle sediment. There are three authigenic minerals: pyrite, dolomite and quartz. Pyrite is present mainly as irregularly shaped grains with scattered cubic and octahedral forms, and as rare, minute spherical grains. Small pyrite grains are included in some dolomite rhombs and secondary quartz, and pyritic films partly outline the cores of some of the secondarily enlarged quartz grains. Pyrite therefore preceded both minerals. Many small dolomite rhombs are completely enclosed by secondary quartz, but it is notable that dolomite never penetrates the detrital quartz cores. Where enclosed dolomite impinges against detrital quartz, the shape of the contact is governed entirely by the shape of the original quartz grain (see Figs 4, 5e and 5f). The relationships conclusively show that dolomite grew before authigenic quartz. Therefore the order of formation of the diagenetic minerals is: pyrite first, dolomite second and quartz third. A remarkable feature of some dolomite boundaries is that they are parallel to the margins of clastic grains, but are separated from them by a thin film of authigenic quartz. This shows that some dolomite grains have been slightly replaced by the last-formed mineral, quartz.

Septimus Limestone.—Textures that are fairly widespread in Western Australian Palaeozoic limestones are found in a calcareous sandy dolomitic rock from the Carboniferous Septimus Limestone in the Bonaparte Gulf Basin. The rock studied (No. 36879, from the foot of Mt. Septimus on the northern side) is slightly weathered and is made up mainly of sparry calcite (6%), quartz (40%), and dolomite (53%). The dolomite has a striking appearance as it contains large, brown, opaque limonitic zones which represent the light-grey zones of clay-sized impurities found in unweathered rocks of the formation (Plate III, Fig. 3). Brachiopod and crinoid fragments are abundant elsewhere in the formation but are sparse in this specimen. Many of the quartz grains were initially well rounded but are now angular from enlargement. The environment of deposition seems to have been neritic, and there are three diagenetic changes of importance; namely formation of the zoned dolomite, formation of the sparry calcite, and growth of the quartz.

Zoned dolomite crystals have been recorded elsewhere by numerous workers, including Gilbert (*in* Williams, Turner and Gilbert 1954, pp. 350-351), Pettijohn (1957, pp. 422-3) and Taft (1961). The last-named author describes dolomite, probably forming at present near the sediment-water interface in Florida, which contains dark cores apparently of organic matter. It does not therefore seem necessary to postulate two periods of growth to account for the dark and clear zones in dolomite of the Septimus Limestone, for by analogy with the Florida dolomite the mineral is most likely to have formed in a marl or impure calcilutite matrix, the impurities of which were arranged to form the dark zones. The patches of sparry calcite certainly represent a recrystallized matrix for they are too large to have been precipitated in pores between clastic crystals, and furthermore, they do not have the micro-drusy texture characteristic of calcite precipitated in pores. Non-calcareous material must therefore have been expelled during recrystallization. There is other evidence to support the view that the dark zones of the dolomite are impurities derived from a matrix which has since recrystallized and expelled non-calcareous material. Dolomite crystals that replace clay-sized matrices rarely have a preferred orientation, whereas dolomite that replaces calcite crystals, such as crinoid fragments, generally adopts the same crystallographic orientation as the replaced calcite crystal. In this rock numbers of differently oriented, zoned dolomite crystals are found in the one sparry calcite crystal, indicating replacement of a fine matrix before its recrystallization (see Plate III, Fig. 3).

Dolomite formed before the quartz outgrowths, for although it is commonly partly surrounded by secondary quartz, it never penetrates clastic quartz grains when it abuts them. However, the relationship between the dolomite and quartz is more complex than indicated, for the secondary quartz has partly replaced some dolomite, leaving irregular boundaries between them, but has not attacked other dolomite, so that their boundaries are planar. All planar boundaries between secondary quartz and dolomite are dolomite faces. On the other hand, all planes developed between quartz and calcite are quartz faces. This suggests, but does not prove, that the secondary quartz grew after the dolomite but before transformation of the matrix to sparry calcite.

It is worth noting that where adjacent calcite and dolomite are replaced by the same quartz outgrowth, the quartz commonly penetrates more deeply into the calcite than into the clear marginal parts of the dolomite crystal. This, and the fact that some dolomite is enclosed but not replaced by quartz, indicates that it was less prone to replacement by quartz than the calcitic matrix.

The diagenetic history of the rock, as tentatively deduced from its textures, can now be summarized. The original sediment is believed to have been a shelly and sandy impure calcareous mud (or marl.) The first mineral to form was dolomite, which partly replaced the matrix, perhaps before significant consolidation, and arranged included non-calcareous material in zones. Then the quartz grains were enlarged, the secondary quartz partly surrounding dolomite crystals and partly replacing them, but evidently replacing the adjacent calcitic matrix more easily. Finally, perhaps because of the pressure of compaction, the remaining matrix was converted to coarse crystalline calcite, and impurities were dissolved and squeezed out.

Miscellaneous Silicified Limestones.—A detailed description of a partly silicified limestone, the Sakmarian Woolaga Limestone Member of the Holmwood Shale, in the Irwin River Basin, has been given by G. Playford (1958, pp. 53-62). Playford describes a diagenetic sequence that includes intrastratal solution (producing vugs), reorganization of calcite, and finally, partial replacement by chalcedony.

The best known sedimentary sequence in Western Australia whose lithology is ascribed to silicification of limestone is that of the Coomberdale Chert, a formation over 3,300 feet thick within the Moora Group. The chert has been described by Logan and Chase (1961) as a novaculite made up of microcrystalline quartz, with only rare chalcedony. They cite considerable evidence to support their theory of replacement, including the observed transition of dolomitic limestone to chert, the presence of siliceous fossils and ooliths believed to have been originally calcareous, relict carbonate inclusions in the chert, siliceous rhombs apparently after dolomite, and quartz overgrowths on detrital quartz in some clastic members. Some of the original limestone is silicified only along joints, showing that the process occurred after lithification.

Many American and European limestones and dolomites yield, on solution, residues of fine-grained, euhedral authigenic quartz and feldspar, generally without recognizable clastic cores. Mere cataloguing of the literature on these minerals, which were the first to be recognized as authigenic, would unduly lengthen this paper, but most early references are in Boswell (1933). Authigenic quartz, feldspar and tourmaline have been recorded from dolomite in the Gap Creek Formation (Glover 1955, p. 3) but generally there has been little attempt at extraction of such residues from other Western Australian limestones.

De-silicated Rocks

The replacement of carbonate minerals by various forms of silica is not a one-way process, and it is probably at least as usual for carbonate to replace silica. Penetration of rounded quartz grains by the calcareous matrix of sandstones is an unmistakeable and rather common indication of this type of replacement (Plate III, Fig. 4) and it is not unusual for the corroding calcite to form deep re-entrants or even to split the quartz grain in two. In Western Australia, clear examples of these replacement textures are found in parts of the Pleistocene Coastal Limestone [see Glover *in* Ride *et al.* (1962, p. 23); and Hodgson *in* Hodgson *et al.* (1962)]. Hodgson has also described rocks from the Cheyne Bay area in which quartz grains are corroded by opal, limonite and goethite, and he suggests that these minerals replaced a calcilutite matrix that was initially responsible for the corrosion. Although the formation of iron oxide cements may generally be a function of weathering rather than diagenesis, if it causes quartz to dissolve and be available for later precipitation perhaps at considerable depth, it clearly has diagenetic significance. Many ferruginous sandstones in Western Australia contain corroded quartz grains, as for example those described by McLellan (*in* Brien and McLellan 1962) from the Cockleshell Gully sandstone, but there has not been enough work to justify generalization about the role of the ferruginous cement.

Miscellaneous Diagenetic Minerals and Textures

Glauconite.—The mineral glauconite, a product of early marine diagenesis, which has been mentioned in the discussion of the Clark Sandstone, has been recorded from the following Western Australian units: Clark Sandstone (Cambrian); Emanuel Formation, Pander Greensand (Ordovician); Blina Shale (Triassic); Langey Siltstone, Callawa Formation (Jurassic); Birdrong Formation, Muderong Shale, Windalia Radiolarite, Alinga Formation, Toolonga Calcilutite, Pcepingee Greensand, Molecap Greensand, Gingin Chalk, Poison Hill Greensand, South Perth Formation, Madura Shale (Cretaceous); Boongerooda Greensand, Wadera Calcarenite (Palaeocene); Jubilee Calcarenite, Giralia Calcarenite, Osborne Formation, Plantagenet Beds (Eocene). Glauconite is present in other formations, and the known number of occurrences will probably increase with further drilling. Veevers and Wells (1961 p. 146) point out that glauconite is found in practically all subsurface occurrences of Mesozoic finer grained rocks in the Canning Basin, and its absence in outcrops is almost certainly due to removal by weathering. Hodgson (1962) has discussed the origin of the glauconite in the Plantagenet Beds at Cheyne Bay and concludes that much of it has been derived from mica.

Phosphatic Minerals.—Phosphatic nodules have been recorded from several Western Australian units, namely Bogadi Greywacke (Permian); Colalura Sandstone, Bringo Shale (Jurassic); Molecap Greensand, Poison Hill Greensand (Cretaceous). The phosphatic nodules in the Colalura Sandstone and Bringo Shale have been described by P. E. Playford (1959) who considers that they grew before burial in the sediment, possibly by growth from coprolitic cores. Fossil wood replaced by collophane occurs in the Colalura Sandstone.

The best known phosphate deposits are near Dandaragan in the Molecap Greensand, where calcium phosphate (collophane) nodules are also associated with phosphatized wood. Other phosphatic minerals present in places in the matrix, and apparently derived from collophane, are dufrenite, vivianite, beraunite, wavellite and minylite (Matheson 1948). Matheson believed that the phosphatic material was initially dissolved from organic remains and deposited as collophane nodules during sedimentation and for some time after, before consolidation of the sediments.

A pale to dark-brown mineral that replaces shell fragments in the Pleistocene Coastal Limestone was tentatively identified as collophane by Hodgson (*in* Hodgson *et al.* 1962), and its identification has been confirmed by chemical tests (Hodgson, personal communication). The mineral is very widely but sparsely distributed in the formation, and has been seen by the present author from Shark Bay and Rottnest Island, generally comprising less than one per cent. of the rock. It is possible that during the Pleistocene there were numerous widely scattered and short-lived pockets of very high organic activity, in which the shells became phosphatized. The eustatic fluctuations of the Pleistocene would have depleted populations of these localized environments only to allow them to become strongly established elsewhere. The phosphatized shells would be incorporated in the repeatedly reworked terrigenous and organogenic detritus, and thus attain their present wide but sparse distribution.

Gearksutite.—The rare mineral gearksutite, a hydrous fluoride of aluminium and calcium, has been described from a phosphatic layer near the base of the Cretaceous Poison Hill Greensand, about two miles east of Gingin, by Simpson (1920). He ascribes its formation to growth *in situ*, from interaction between solutions of the nearby minerals fluorapatite (present as nodules) and gibbsite (a constituent of overlying laterite). The reaction postulated by Simpson seems to be on the borderline of late diagenesis and weathering.

Jarosite.—The secondary mineral jarosite has been noted by Prider (1943, p. 39) in the Precambrian Cardup Shale and by Clarke *et al.* (1951) in Permian formations of the Irwin River area. The latter authors believed that euxenic conditions led to formation of pyrite or marcasite, from which the jarosite was later developed. This view accords with the origin suggested for jarosite in Mexico by Pough (1941), and in Yorkshire by Hartley (1957). As with gearksutite, its formation is due to either late diagenesis or weathering, depending on arbitrary definition.

Tourmaline.—Many quartz sandstones contain rare but widely distributed grains of clastic tourmaline showing authigenic outgrowths. Clarke *et al.* (1954, p. 20) have described authigenic elbaite on rounded schorlite grains from quartzites of the Precambrian Mt. Barren Group, but it should be noted that these rocks have been metamorphozed. Prider (1943, p. 41) has described tourmaline euhedra from the Precambrian Cardup Shale, but again there is doubt about their origin, and they may be metasomatic.

Concretions.—Concretions, mainly calcite and limonite, are common in many Western Australian sediments, but they have generally not been the subject of detailed studies. Calcareous and sideritic concretions from the Learmonth Formation, Carnarvon Basin, have been described, and evidence for their formation before much compaction of the rock has been cited (Glover 1960*a*). Barite concretions are known from the Wogatti Sandstone and Gearle Siltstone of the Carnarvon Basin. The varied concretions in sedimentary rocks of the Irwin River area have been suggested as a subject of research by Clarke *et al.* (1951, p. 47).

Kaolinite.—Crystalline kaolinite, apparently of authigenic origin, is found in sandy claystone of the Wagina Sandstone, 13 miles east-northeast of Mingenew. The specimen examined with the universal stage consists of a clay mineral (80%), angular quartz grains (18%), opaque material (2%) and muscovite and biotite (<1%). The clay mineral, which was shown to be kaolinite by differential thermal analysis, is present in two forms. Most of the clay consists of a brown paste, but scattered through it are pale-brown, anhedral to subhedral crystals about 0.3 mm long. These crystals have uneven extinction, and their main cleavage is generally parallel to the bedding, but some are bent and have a vermicular form. A few grains, when cut parallel to their cleavage, show pseudo-hexagonal habit due to lines of minute inclusions arranged at 60° (Plate III, Fig. 5). $2V_\alpha$ ranges from 7° to 40° (20 readings), and most values are between 15° and 30°, a little lower than those generally quoted for the mineral.

Kaolinite is too soft to survive transport as large crystals, and these must have grown in the sedimentary rock. Post-compactional growth due to percolation of solutions can be ruled out, as the rock is impervious. The grains therefore probably grew before compaction, or during compaction as fluids were squeezed out.

Allophanoids.—A white, isotropic, tubcrose mineral, probably an allophanoid, has been recorded by Prider (1948, pp. 112-113) as an authigenic constituent in the finer grades of the ferruginous sandstones of probable late Mesozoic age at Ridge Hill.

Outstanding Causes of Diagenesis

Reactions Near the Sediment-water Interface

It is generally accepted that much glauconite and collophane have formed by reactions on or just below the sediment-water interface, and the evidence has been summarized in Carozzi (1960) and other recent texts. This origin accords well with the appearance of the glauconite and collophane seen in this investigation, and with their relationship to other minerals. Pyrite is another mineral that is commonly of early diagenetic origin, and it is found in many rocks containing glauconite and collophane. It has long been suspected that much of the pyrite of black organic shales has been precipitated by sulphate-reducing bacteria, and the process has been described recently for sediments on the McMurdo Sound region of Antarctica by Barghoorn and Nichols (1961). Love (1958) believed that residues left after solution in nitric acid of framboidal pyrite from the Lower Carboniferous Oil Shale of Scotland, were the remains of such bacteria. The present writer obtained similar residues from the Jurassic Learmonth Formation of Western Australia, but described them as remains of organic fragments including spores, which were etched and frayed (but not much compressed) during pyritization. The spores and other organic fragments were thought to have provided micro-environments of putrefaction in which the pyrite grew by bacterial action (Glover 1960*a*).

Pyrite, as minute spherical and subspherical grains and as crystals (commonly cubes and octahedra) often occupies the cells of wood fragments, and has evidently formed before compaction of the wood. It is found in some calcareous and sideritic concretions that also contain virtually uncompressed fossils and must themselves have formed early. It is frequently included in dolomite rhombs and in quartz outgrowths, clear evidence that it formed before them. In ancient sediments, therefore, pyrite is often demonstrably one of the first products of diagenesis, and its common association with organic detritus accords with bacterial origin in a reducing environment.

Ricour (1960) has drawn attention to the possibility of bacterial origin for dolomite, and that mineral is certainly the first diagenetic product to have formed after pyrite in the Western Australian sediments examined, and it could therefore have grown in reducing muds. There is no doubt that dolomite has formed before the sparry calcite matrix of some limestone (see the description of Septimus Limestone), but its origin warrants fuller comment, and will be considered later.

Limestones commonly contain authigenic quartz and feldspar euhedra (that is, completely authigenic grains as distinct from the secondary outgrowths on clastic cores found in many sandstones), and many early workers attributed them to growth in the original calcareous mud. There are generally no significant clues to indicate whether growth was early or late. However, there are rare records of authigenic feldspars in shales and siltstones (Gruner and Thiel 1937) and it is hard to see how these feldspars could have formed after lithification, for the rocks would presumably have been impervious to solutions that might have precipitated them.

Much of the literature on diagenesis, particularly early diagenesis, assumes that it is promoted or assisted by the presence of organic matter. Sujkowski (1958, p. 2694) emphasized this view in his discussion of the agents of diagenesis and stated "Water is the main agent of diagenesis and organic matter is an auxiliary."

Compaction

Many mineralogic changes in fine-grained sediments such as claystones and shales, and in sediments with an abundant clay-sized fraction, such as greywackes, seem to have been caused at least partly by compaction. These changes includes the formation of micas, chlorites and coarse kaolinite crystals in claystones and shales, and the growth of chlorites and other flakey minerals that penetrate sand grains (with corresponding solution of quartz) in greywackes. It is unlikely, in view of the impermeability of these rocks, that solutions could have circulated in them after lithification. Lutites compact most rapidly during and shortly after deposition (Jones 1944) and although there may be slow compaction during deepening burial, the abundant liquids squeezed through the compacting rocks in the early stages probably assist in the changes listed above.

The formation of sparry calcite from calcilutite, apparently caused by compaction in some limestones, is commonly accompanied by expulsion of carbonaceous matter, clay minerals and other finely divided impurities. This can be demonstrated in dolomitic limestones where dolomite crystals have replaced some of the matrix and are grey from zones of included impurities, whereas later recrystallization of the surrounding matrix has left it clear and sparry.

Other changes generally ascribed to compaction include development of macro- and micro-stylolites in limestones (Stockdale 1922, Dunnington 1954).

Waldschmidt (1941) suggested that pressure between quartz grains of sandstones at their points of contact leads to solution, development of micro-stylolitic contacts and precipitation of the quartz nearby on grain boundaries where pressure is less. This is probably responsible for some redistribution of quartz in sandstones, but as Pettijohn (1957, p. 657) points out, many micro-stylolitic contacts in sandstones are between the enlarged grains rather than adjacent detrital areas. Pettijohn earlier (1949, p. 481) attributed all well-developed sutures between enlarged grains to orogenic deformation after authigenic growth, and all examples so far observed by the present writer have been in deformed rocks.

Purely physical effects of compaction are seen in the bending of micas and chlorites around sand grains, and in the flattening of some fossil remains.

Intrastratal Solution and Precipitation.

The main causes of diagenesis in arenites are solution and precipitation by fluids percolating between the grains. The chief fluid by virtue of its ubiquity, is undoubtedly water, but the possible diagenetic effects of oil have provoked little attention. Kulbicki and Millot (1960) consider the role of oil in some early Palaeozoic Saharan sandstones, and conclude that it has preserved kaolinite from the effects of salt waters, which elsewhere in the sequence have altered it to illite.

Solution.—Textural evidence of solution is afforded by the raggedness of many heavy minerals, particularly ferromagnesian silicates, and this is generally best observed in grain mounts of heavy minerals concentrated from disaggregated sedimentary rocks. The susceptibility to solution of many heavy minerals has long been recognized, notably by Bramlette (1929, 1941), Edelman and Doeglas (1932), and Smithson (1941). Pettijohn (1957) has summed up much of the literature on the subject. However, not only the heavy minerals are affected. Textures resulting from the solution of quartz and feldspar grains, and of lithic fragments, are readily evident in thin sections of many Western Australian sandstones. The grains show ragged boundaries, re-entrants and even skeletal textures, and are mostly replaced by carbonate or ferruginous cement. Many examples of corrosion of sand grains by diagenesis have been reported in the recent literature, and in fact it is now appreciated that surface textures such as frosting and pitting may be due to shallow corrosion as well as to abrasion*. A significant feature of this corrosion is that it provides an apparent source for much of the material that elsewhere forms the authigenic outgrowths on quartz, feldspar, and other minerals.

* Strangely enough, frosting is said to be produced both by etching (Walker 1957) and by deposition of secondary silica (Humphries 1961).

PLATE I

Fig. 1.—Barite-cemented quartz sandstone from the Jurassic Wogatti Formation (Rough Range No. 1 Bore, core 9, 3,738-3,752 ft), as seen with the universal stage. The three quartz grains (light grey, no cleavage) appear to have irregular boundaries with the barite (dark grey, cleavage). Plane polarized light. Width of field 0.4 mm.

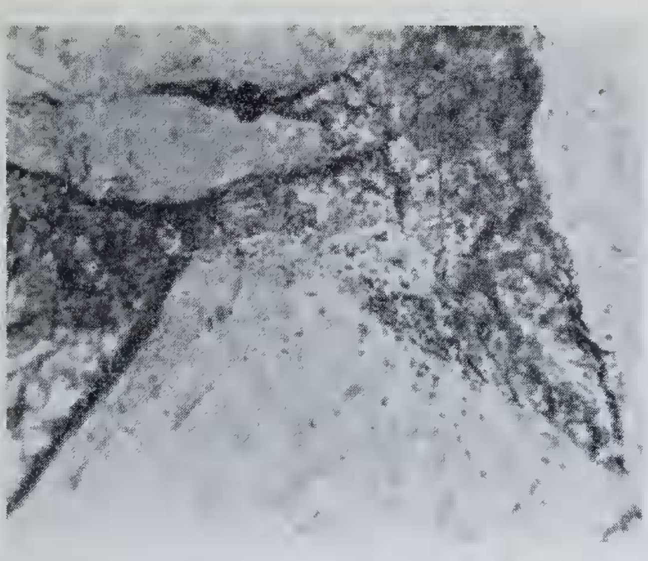
Fig. 2.—Same field as in Fig. 1 after tilt of the stage through 25°. An authigenic quartz outgrowth with at least three faces is evident, and other faces can be detected with different tilt. Plane polarized light.

Fig. 3.—Glauconite and quartz in the Clark Sandstone (No. 48629). Authigenic quartz outgrowths tend to be moulded by the glauconite pellet, which is crossed by a small crack or micro-fault. Glauconite to the right of the crack is securely wedged between two quartz grains, but below the crack it has been displaced about 0.02 mm, apparently by growth of the quartz grain in the lower left of the field. Plane polarized light. Width of field 0.6 mm.

Fig. 4.—Glauconite and K-feldspar in the Clark Sandstone (No. 48629). The core of the feldspar, which cannot be sharply distinguished from the authigenic rim in the photograph is microcline. The rim has patchy extinction with 2V negative, about 64°, and is adularia. 001 cleavage (parallel to the well-developed faces) and 110 cleavage pass from core to rim. The authigenic feldspar is moulded against the glauconite, and its growth has apparently caused the glauconite to yield along a small fault. Plane polarized light. Width of field 0 6 mm.

Fig. 5.—Microline with an authigenic rim in the Clark Sandstone (No. 48629). The rim has slightly uneven extinction, is crowded with haematite inclusions, and has optics close to sanidine. In the rim 2V ranges, in different places, from 41° to 18° (approx.). 001 and 010 cleavages are present in the core, and the 001 cleavage can be seen passing from the core to the rim. Plane polarized light. Width of field 0.35 mm.

Fig. 6.—Same feldspar grain as in Fig. 5 under crossed nicols. Note how the wedge-like albite twins of the microcline core stop at the margins of the authigenic feldspar. Other grains are glauconite (right), feldspar (upper left) and quartz.



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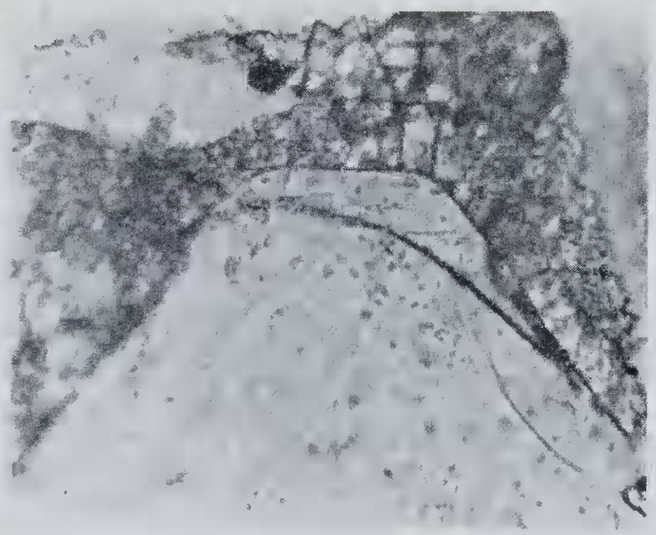

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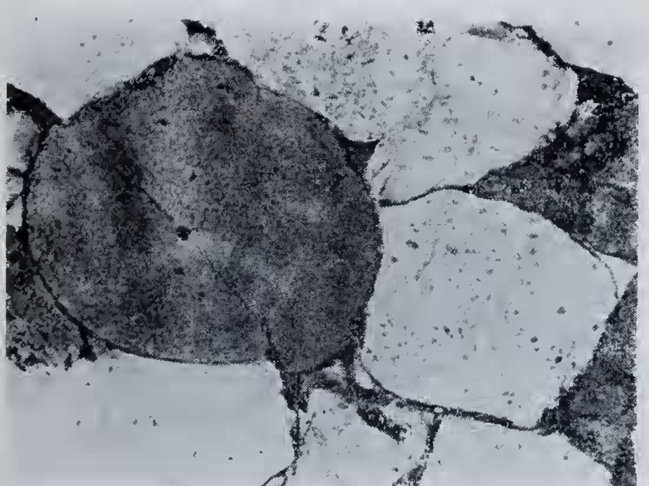

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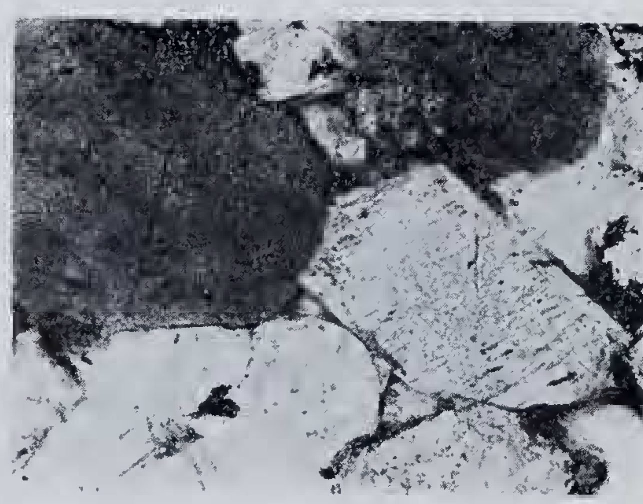

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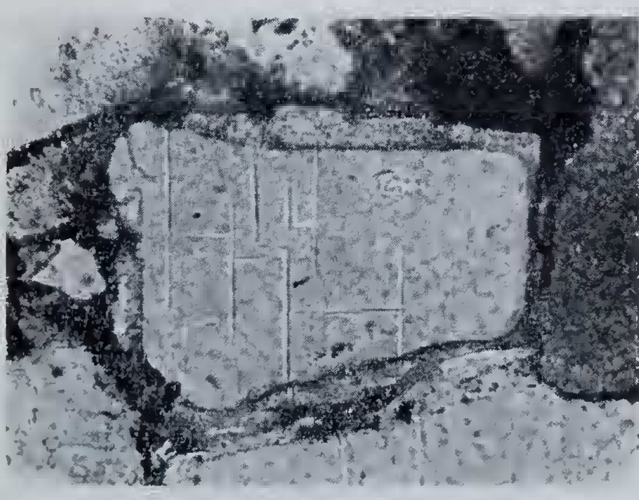

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PLATE I

Precipitation.—Precipitation from intrastratal solutions is a major factor in the diagenesis of arenites, and accounts for most outgrowths on quartz and feldspar. Pressure-solution of grains at their contacts, and precipitation nearby on grain boundaries where pressure is less, is probably responsible for some redistribution of quartz. However, pressure-solution is only a minor feature of the texture of most West Australian sandstones examined, and the process seems inadequate to account for the abundant secondary quartz in many of them. Moreover, some silicified sandstones show no pressure-solution effects and it can only be concluded that the quartz grains have acted as seed crystals and have been enlarged by percolating siliceous solutions. The same process is responsible for enlargement of feldspar grains.

Naturally, sand grains do not act as seed crystals if they are not themselves crystals, but are fragments of opaline or chalcedonic chert, schist, slate, volcanic glass or fine-grained volcanic rock, or if they are pellets and ooliths composed of clay-sized calcareous material. Shell fragments (except echinoderms, which break into coarsely crystalline debris) do not act as seeds. Therefore in lithic sandstones and calcarenites many pores between clastic grains can not be filled with minerals that are optically continuous with the clastic fragments. These pores instead are commonly filled with prismatic or fibrous minerals with their length toward the centre of the cavity, or else they are lined with several concentric mineral bands. Both of these textures could probably arise only by precipitation from circulating solutions. Finally, a significant point of negative evidence favouring precipitation from circulating solutions is the general absence of enlarged grains in greywackes and sandy lutites, where post-compactional circulation of fluids is practically impossible. These rocks often contain corroded quartz and feldspar grains, and the dissolved material was probably removed by the solutions squeezed out during compaction, to be made available for precipitation elsewhere.

One of the problems connected with the precipitated material is whether or not it normally travels a long way before deposition. There is evidence that it often does not, for it is characteristic of many authigenic minerals that they are most abundant in those sediments which are able to supply their component elements. Thus it was pointed out by Gilbert (*in* Williams, Turner and Gilbert 1954) that quartz cement is abundant only in quartz-rich sandstones, authigenic feldspars are found chiefly where there is also detrital feldspar, and carbonate cements, although found in all types of sandstones, are invariably formed wherever the original sand contained primary carbonate. He also points out that laumontite and chlorite are generally found in sandstones containing volcanic fragments and are much less abundant in others, and lists several Tertiary Californian sandstones with volcanic detritus and chlorite cement. Montmorillonoid, a mineral that resembles chlorite in containing magnesium, has been described coating grains in late Tertiary lithic (volcanic) sandstones in California by Lerbeckmo (1957), and authigenic chlorite is described in similar sandstones from Mazakhstan by Chernikov (1960). Australian chloritic sandstones with volcanic detritus include the Arrowsmith Sandstone (pp. 37-38) and numerous Permian and Triassic sandstones from the Bowen Basin, Queensland. A local exception to the generalization seems to be the Poole Sandstone of the Canning Basin, Western Australia, which is a feldspathic sandstone with chlorite shells around the grains, but apparently no volcanic material (Johnson and Dallwitz *in* Veevers and Wells 1961, p. 267). As Gilbert also noted,

PLATE II

Fig. 1.—Slightly kaolinized arkose from the Mokadine Formation (No. 36909). All grains except for opaques and the three clear quartz grains at bottom left are K-feldspar. Authigenic quartz and feldspar outgrowths have practically eliminated pore space. None of the boundaries of authigenic material in this field corresponds precisely with a common crystal face. A quartz grain has slightly penetrated the long feldspar grain, and the clear microcline (centre, right) has slightly penetrated the kaolinized adularia below it. The penetrations were probably caused by compaction before authigenesis. Plane polarized light. Width of field 0.6 mm.

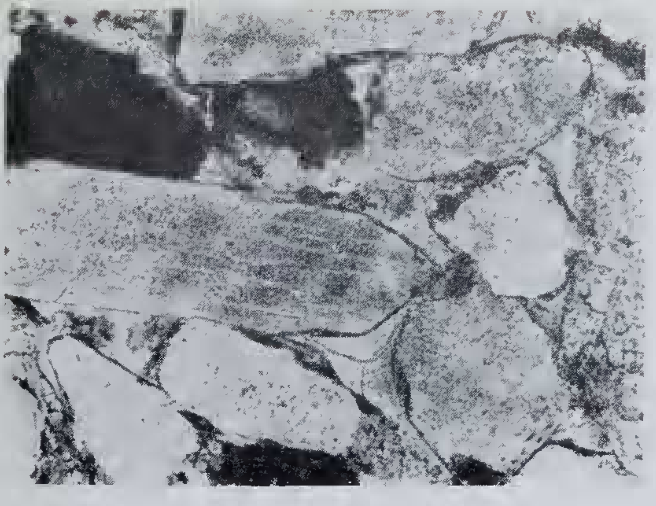
Fig. 2.—Micro-drusy texture in the Arrowsmith Sandstone (No. Pf2). Clastic grains are K-feldspar (left, with cleavage) and quartz. The cavity between them has been filled by three minerals, which are, from the outside: (1) a thin film of brown mica-like mineral, (2) a pale green chlorite layer with minute serrations directed toward the centre of the cavity, and (3) an infilling of orthoclase ($2V_\alpha = 43°$). A little chlorite, probably projecting from a surface just grazed by the section, can be seen apparently within the orthoclase. Width of field 0.3 mm. Plane polarized light.

Fig. 3.—Micro-drusy texture in the Arrowsmith Sandstone (No. Pf2). The cavity on the right has been filled with three minerals which are, from the outside: (1) a thin film of brown mica-like mineral, (2) a pale green chlorite with minute serrations directed toward the centre of the cavity, and (3) an infilling of adularia ($2V_\alpha = 55°$). The sequence is different in the cavity on the left and is, from the outside: (1) the mica-like mineral, (2) quartz (visible only on the lower right of the cavity), (3) chlorite, and (4) quartz, optically continuous with (2). Dark clastic grains are iron-stained volcanic fragments, and others are quartz and feldspar. Plane polarized light. Width of field 0.6 mm.

Fig. 4.—Authigenic enlargement of plagioclase (An_5) in the Arrowsmith Sandstone (No. Pf61). Twinning continuous without interruption into the rim (lower right). The grey mineral is calcite: clastic fragments include quartz, K-feldspar and opaques. Crossed nicols. Width of field 0.9 mm.

Fig. 5.—Micro-drusy texture in Enokurra Sandstone (No. 38725). A fringe of sericite whose minute flakes point toward the centre of the cavity, adheres to the iron-stained surfaces of the clastic grains. This was the first authigenic mineral to form. The rest of the cavity is filled with quartz, most of it optically continuous with the clastic grain in extinction (lower left). Crossed nicols. Width of field 0.3 mm.

Fig. 6.—Authigenic quartz in the Cockatoo Sandstone (No. 48624). Most of the pore space has been filled. Some boundaries correspond to common crystal faces, but many do not, and some are curved. The clastic grains are outlined by a film of haematite. Plane polarized light. Width of field 1 mm.



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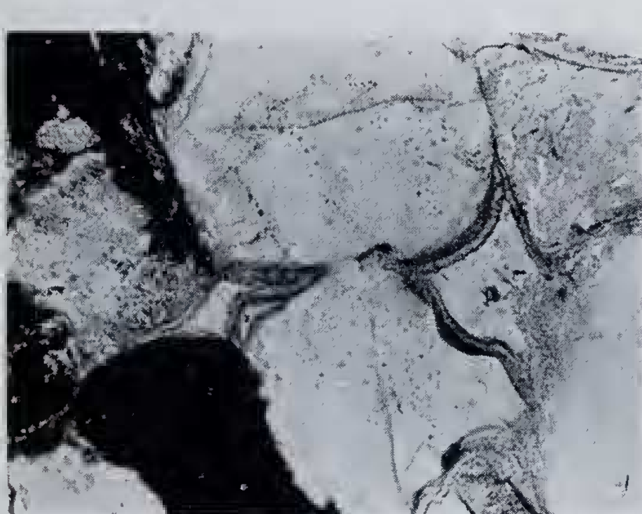

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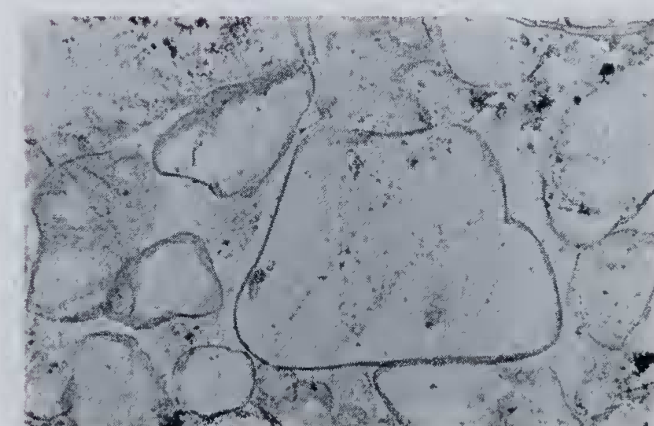

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PLATE II.

part of the tendency for particular cements to concentrate in certain sandstones is due to the presence of sand grains which act as crystal nuclei on which cement of the same composition is precipitated. This holds with quartz and feldspar, but not with chlorite, so the presence of the latter in sandstones with volcanic material is generally convincing evidence for its local derivation. It must clearly be understood that the preceding discussion applies to precipitated chlorite, and not to that which results from reorganization of the finely divided matrix in rocks such as greywackes and shales.

The replacement of carbonate in many limestones by quartz, chalcedony, opal and feldspar seems contrary to the general principle of intrastratal (as opposed to interstratal) origin for the solutions. There can be no doubt that many such solutions have originated elsewhere in more siliceous strata, but the replacement process is sufficiently complex to warrant brief separate treatment below.

The Carbonate-silica Relationship.—Silicification of carbonates has to be considered with its reverse process, the equally common carbonation of silica and silicates in calcareous sandstones and other rocks. The considerable literature on the chert problem, particularly as it applies to the distinction between primary and secondary chert, is well summarized in Pettijohn (1957), and will not be reviewed here. Some of the views about the causes of replacement, where it can be demonstrated, will, however, be mentioned.

Much of the discussion about replacement has hinged on the chemistry of calcite and amorphous silica, particularly on their solubility under different pH conditions. It has been shown that at 25°C the solubility of amorphous silica is practically independent of pH except where it exceeds pH9, when the silica becomes increasingly soluble with increase in alkalinity (see Krauskopf 1959). On the other hand, calcite becomes less soluble with increase in pH.

Recent papers to use these data include that of Rouge *et al.* (1959) who explained a decrease in quartz and silicates of certain limestones as they became more strongly dolomitized by assuming that the quartz and silicates dissolved because of the increased pH that caused dolomitization. Examples of chert replacing carbonate and vice versa in the same rocks have been cited by Walker (1962) and ascribed either to variations in pH where the interstitial waters are highly alkaline ($pH > 9$) or to variations in temperature attendant on deep burial. The prospect of sufficiently high pH values for these reactions in the natural environment has also been accepted by Dapples (1962, p. 917), but not by Siever (1962) who has discounted the significance of pH, and has put forward the idea that clays act as semi-permeable membranes causing solution of carbonate and precipitation of silica. He has also appealed to several other mechanisms.

Generalizations about the distribution of authigenic quartz, chalcedony and opal are possible, though much is still unknown about their origin. Quartz forms very commonly as outgrowths on clastic quartz nuclei in many rocks including calcareous sandstones and sandy limestones, and these outgrowths are generally fairly free of inclusions. Quartz also grows as minute, widely distributed euhedra without obvious clastic cores in some limestones, but the euhedra are often crowded with carbonate inclusions.* Authigenic feldspar assumes both forms, but is far less common. Fine-grained to micro-crystalline quartz forms many cherts, some of which, like the Coomberdale Chert, are considered to be epigenetic. Chalcedony also replaces carbonate widely, but opal generally seems to be Tertiary or younger, and it tends mainly to replace clay-sized carbonate, though it is not restricted to that lithology. Most pre-Tertiary opal has probably crystallized to micro-crystalline and fine-grained quartz. Selective replacement by different forms of silica often leaves precise details of the original texture as in Plate III, Fig. 6. It has been assumed by Millot *et al.* (1959) that percolating waters with low concentrations of dissolved silica cause precipitation of quartz, whereas solutions with high concentrations, and often many impurities, result in opal and chalcedony. However, the

* Although quartz and feldspar euhedra in limestones were the first generally recognized products of authigenesis they have yielded practically no textural evidence of their time of formation, variously suggested as very early to late in the history of the limestone.

PLATE III

Fig. 1.—Micro-drusy texture in the Coastal Limestone (No. 50267). Calcite fibres partly cement a coarse-grained calcarenite. Plane polarized light. Width of field 0.8 mm.

Fig. 2.—Micro-drusy texture in an oolite from Devonian limestones of the Lennard Shelf (No. 50269). Sparry calcite completely fills the voids between individual ooliths. Plane polarized light. Width of field 1.75 mm.

Fig. 3.—Dolomite and sparry calcite in the Septimus Limestone (No. 36879). There are a few quartz grains in the top right corner. All calcite is part of the one crystal (see cleavage). Slight weathering has oxidized iron and emphasizes the zone of impurities in each dolomite rhomb. The dolomite crystals show no preferred orientation, and apparently grew in an impure calcareous mud which later crystallized to sparry calcite, expelling impurities. Plane polarized light. Width of field 1 mm.

Fig. 4.—Replacement in Coastal Limestone (No. 8656). Rounded quartz grains are partly replaced by a finely divided clay-calcite matrix. Plane polarized light. Width of field 3.6 mm.

Fig. 5.—Authigenic kaolinite showing pseudo-hexagonal structure due to lines of inclusions, in sandy claystone of the Wagina Sandstone (No. 36927). The matrix is finely divided kaolinite, and there are a few quartz grains. Plane polarized light. Width of field 0.6 mm.

Fig. 6.—Silicification in the Toolonga Calcilutite (No. 50363, from ¼-mile north-west of Yaringa North Homestead). This silicification is not widespread, but is very selective. Foraminifera are replaced by chalcedonic quartz, and the groundmass is opal with calcareous inclusions. Glauconite (not shown here) is not replaced. Plane polarized light. Width of field 1 mm.

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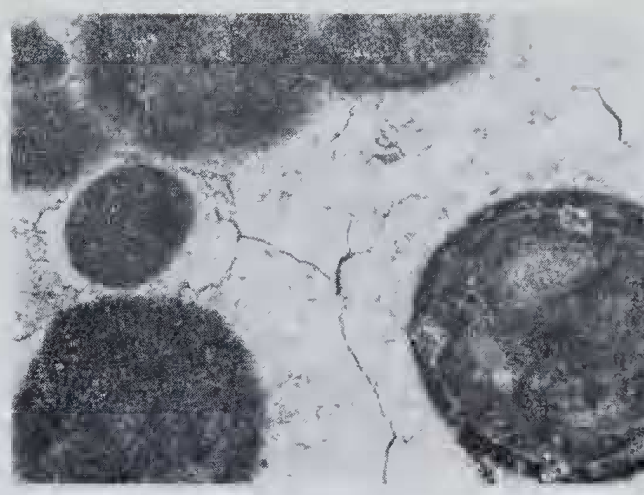

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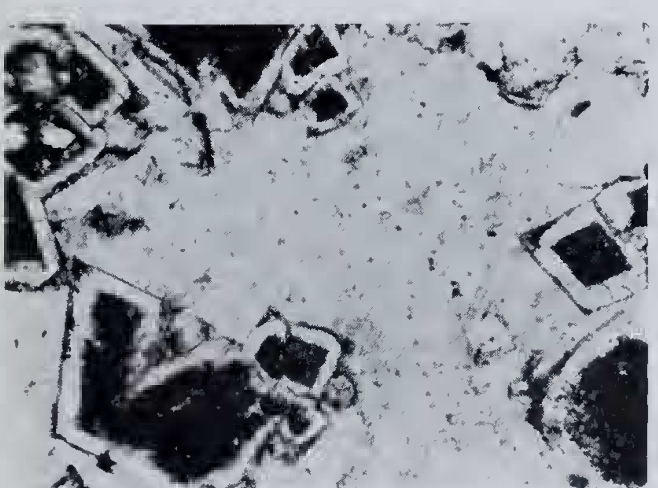

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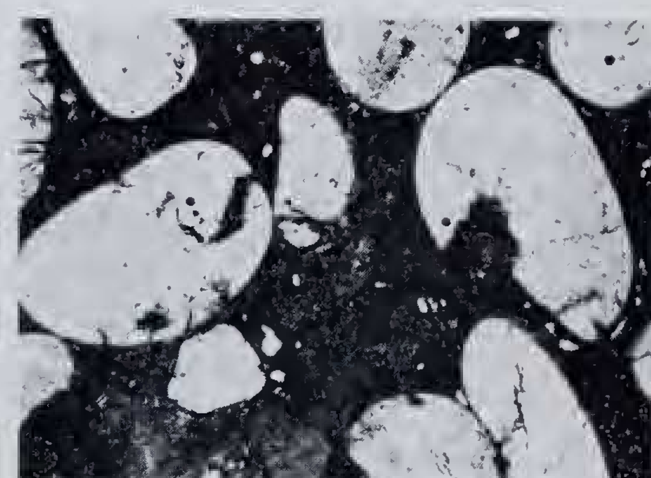

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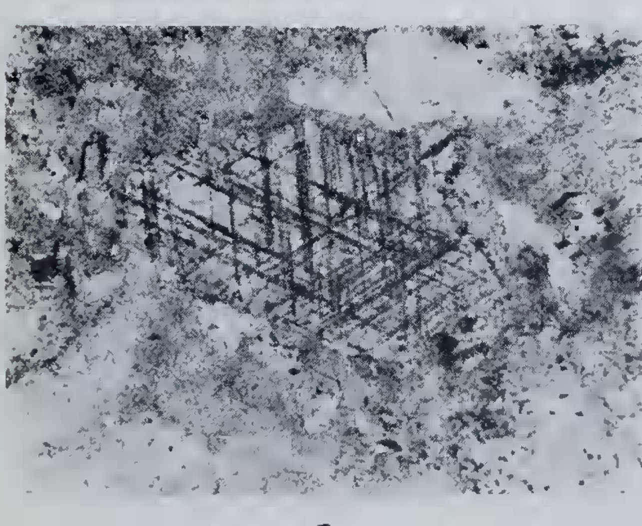

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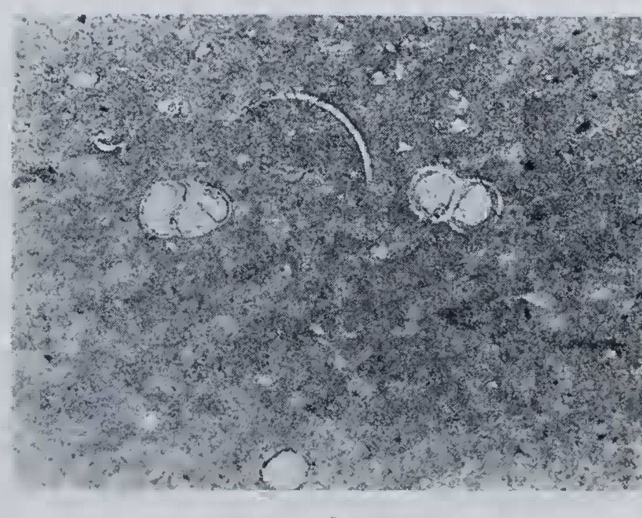

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PLATE III.

tendency observed by Millot (1960, p. 143) for chalcedony to predominate in replacement of carbonate later suggested to him that some intrinsic feature of the calcite itself might influence the type of mineralization. It will be evident from the above summary that much is yet to be determined about carbonate-silica relationships in sedimentary rocks, and that one of the best foundations for advancement is still in the amassing and recording of petrographic data.

The Dolomite Problem

Dolomite is one of the enigmas of sedimentary petrology. It is common in Precambrian and Palaeozoic limestones, less common in Mesozoic limestones, unusual in Tertiary and Quaternary limestones, and apparently forms at present under only exceptional circumstances. Examples of its present day formation are described by Alderman (1959), Miller (1961) and Taft (1961), comprehensive discussions of the origin of the mineral are given by Fairbridge (1957), Pettijohn (1957) and Carozzi (1960), and a recent description of its laboratory synthesis is presented by Siegel (1961).

The texture of many dolomitic rocks points to formation of the dolomite in partly compacted muds below the sediment-water interface, where it is unaffected by wave or current action. The dolomitic grains are euhedral, and commonly enclose pyrite, showing that they form after that early diagenetic mineral. These features are consistent with the general absence of dolomite at present in shallow, unconsolidated marine muds. On the other hand the texture of most dolomitic rocks indicates that the dolomite had finished growing before their final consolidation. For example, dolomite is found in some shales and the impermeability of those rocks seems to preclude precipitation of the mineral from percolating solutions after their final compaction. Furthermore, the textures of many limestones show that dolomite formed before micro-stylolites and before crystallization of sparry calcite from calcilutite, both of which are probably results of intense compaction. It is also probable that the porosity of many dolomitic rocks indicates formation of dolomite before solution of interstitial calcite, for the dolomite seems to have served as the framework from which the calcite was leached. Finally, in this respect, it should be noted that much dolomite has clearly preceded authigenic quartz in some sandstones, although in others the reverse relationship has been established. The above textural features, taken together, would apply very well to dolomite that formed as compaction began, but before those diagenetic features that commonly accompany the last stages of consolidation.

Some dolomite, however, probably forms as a product of late diagenesis in consolidated rocks. Other dolomite undoubtedly forms very early in evaporites. Recently Sabins (1962) has recognised rounded "primary" dolomite in Cretaceous sandstones and states that it formed before burial of the sediment, and hence was subject to abrasion and sorting. Primary dolomite is said by Sabins to be restricted to marine sandstones, and he lists 42 formations containing it.

Fig. 5

Enlargement Textures

Simple Enlargement Textures

(a) Calcarenite showing sparry calcite in crystallographic continuity with crinoid debris. Stippled fragments are calcareous pellets.

(b) Quartz sandstone with quartz outgrowths crystallographically continuous with clastic cores. Note faces. Lightly stippled areas represent pores.

(c) Quartz sandstone with pores completely occupied with secondary quartz. Some outgrowths bounded by plane surfaces, some not. No sutured boundaries.

Indentation Textures

(d) Dolomitic sandy marl in which dolomite is partly surrounded by, or has partly penetrated, quartz outgrowths. Texture does not reveal whether quartz or dolomite grew first.

(e) Dolomitic sandy marl, same as (d), except one of the dolomite grains is moulded onto a clastic quartz core. Dolomite therefore preceded secondary quartz.

Enclosure texture

(f) Dolomitic sandstone with dolomite completely enclosed by quartz. Dolomite therefore formed first, and order is confirmed by moulded dolomite (upper centre). Note how a moulded dolomite has retreated marginally (left centre) due to slight solution during silicification.

Pressure-solution Textures

(g) Quartz sandstone with clastic grains showing sutured boundaries due to compaction, deformation or both. Secondary quartz(s) fills voids, may have come partly or completely from quartz dissolved along sutured contacts.

(h) Calcarenite with micro-stylolite due to compaction. Micro-stylolite outlined by iron-stained argillaceous matter and small quartz grains, both insoluble in the particular conditions of its formation here. Sparse distribution of quartz in rock suggests compaction equivalent to field of view.

(i) Quartzite with sutured boundaries between outgrowths due to deformation after diagenesis. As much a metamorphic as a diagenetic texture.

Micro-drusy Textures

Simple Micro-drusy Textures

(j) Calcarenite partly cemented with fibrous calcite. Fibres are elongated normal to grain boundaries.

(k) Calcarenite completely cemented with sparry calcite. Long axes of calcite crystals are normal to grain boundaries.

(l) Lithic (volcanic) sandstone cemented by fibrous chlorite. Texture basically the same as in (j) and (k).

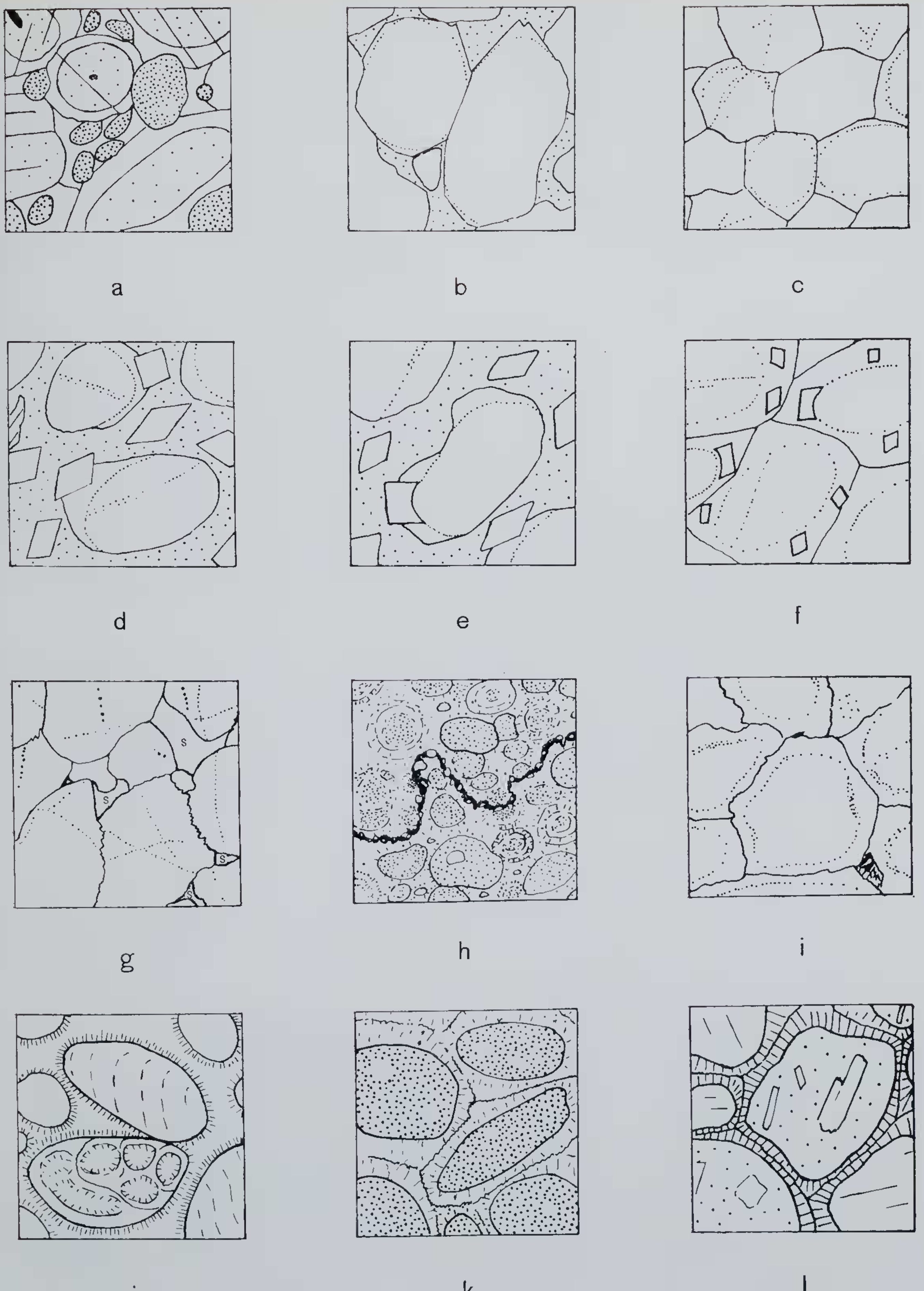

Fig. 5.

To sum up, dolomite almost certainly forms in different ways, and may grow early (by direct precipitation in marine sandstones during sedimentation, or by evaporation) or late (replacing other diagenetic minerals in porous rocks). However, textures in ancient rocks suggest that most dolomite in limestones forms during initial compaction, but before final consolidation. Perhaps the composition of Precambrian and Palaeozoic atmospheres helped to account for its abundance during those eras, as Ronov (1959) believes. A hypothesis which explains the required concentration of magnesia from sea water is that of Adams and Rhodes (1960) who postulate alteration of limestones by magnesia-rich solutions seeping from evaporite lagoons. Further information from the drilling of areas where reefs are now growing seems desirable, though it must be agreed with Ladd *et al.* (1953) that the drilling of atolls has so far yielded puzzling results.

Common Diagenetic Textures and their Interpretation.

Diagenesis, like magmatic crystallization and metamorphism, causes many textures, and, as with those processes, certain basic textures appear again and again. These textures fall in the following categories: enlargement textures, pressure-solution textures, pore-filling or micro-drusy textures, reorganization textures, and replacement textures. Most of them have been encountered in the rocks described earlier, and they are now represented diagrammatically in Figs. 5 and 6. A discussion of their main features and significance follows below.

Enlargement Textures

All enlargement textures are the result of precipitation of minerals on crystal nuclei of the same or similar composition and the best examples are found in porous arenites, where solutions are able to circulate freely. However, some outgrowths have replaced earlier minerals which partly filled the cavities, and these outgrowths may contain clues in the form of minute inclusions.

Simple Enlargement Textures.—Tourmaline and other heavy minerals grow either small projections, or faces, but crystalline calcitic debris, quartz and feldspar usually grow large, simple and well-developed faces, except where prevented by mutual interference (see Figs. 5a, 5b, 5c). The commonly developed faces are, in calcite, rhombohedra: in quartz, prisms and rhombohedra: and in feldspar, basal and side pinacoids and prisms (110 and 130). Faces which are less common in feldspar, but are not rare, include the pyramid (111) and dome (101).

Indentation Textures.—Authigenic outgrowths on one mineral may partly surround, or be partly penetrated, by another authigenic mineral to give an indentation texture. This texture, illustrated in Fig. 5d, does not reveal which of the two minerals grew first. However, in the special case where one of the authigenic minerals has grown so that it is moulded against the clastic core of the other, the moulded mineral has been the first to grow (Fig. 5e).

Enclosure Textures.—These are a development of the indentation texture in which the authigenic rim of one mineral grows sufficiently to include a second authigenic mineral. The included mineral forms first (Fig. 5f).

Pressure-solution Textures

The secondary quartz of some sandstones has been attributed to pressure-solution effects. Where quartz grains impinge on each other under intense pressure, they partly dissolve at points of contact, and interlock with micro-stylolitic boundaries. The dissolved quartz is supposed to precipitate in adjacent pores, where pressure is less. The common association of secondary quartz and micro-stylolitic contacts

Fig. 6

Micro-drusy Textures (continued)

Composite Micro-drusy Textures

(a) Lithic (volcanic) sandstone with pores filled by three minerals which are, from the outside, a micaceous mineral, chlorite (stippled), and feldspar. Note how minute chlorite serrations are directed inward. The micaceous mineral may be a reconstituted clay film on the clastic grains. The order of formation was (1) micaceous mineral (or its precursor), (2) chlorite, (3) feldspar. (Based on a diagenetic sequence in Arrowsmith Sandstone.)

(b) Lithic (volcanic) sandstone showing the following diagenetic sequence: (1) micaceous mineral, (2) quartz, (3) chlorite, (4) quartz (note euhedrism), (5) feldspar. (Based on a diagenetic sequence in Arrowsmith Sandstone.)

Reorganization Textures

(c) Claystone with vermicular kaolinite crystals. Fragility of the crystals is proof of *in situ* formation.

(d) Greywacke with chlorite-sericite matrix, formed from clay-sized detritus. The new flaky minerals penetrate margins of clastic fragments, making this partly a replacement texture also.

(e) Fontainebleau sandstone, in which calcite has reorganized to form large crystals.

Replacement Textures

(f) Calcilutite partly replaced by quartz euhedra with calcareous inclusions. The long fragment is calcite.

(g) Shelly limestone with shells partly replaced by chalcedony, and matrix dolomitized. The dolomite has zonal inclusions—see also (j). Unreplaced shelly material has recrystallized.

(h) Quartz sandstone cemented by sparry barite. The original pyritic and argillaceous matrix is represented by patches of argillaceous impurity, and isolated pyrite grains.

(i) Ferruginous quartz sandstone in which quartz grains are apparently corroded by the ferruginous matrix. One quartz grain shows an outgrowth, product of an earlier diagenetic phase. The texture resembles that where carbonate corrodes quartz, and many such sandstones may be caused by replacement of calcite by iron oxide.

(j) A complex but fairly common texture, in which dolomite has partly replaced the matrix of a sandy marl. Carbonaceous and argillaceous inclusions form a dark zone in each dolomite rhomb. There has been later recrystallization of the matrix to sparry calcite, with expulsion of impurities.

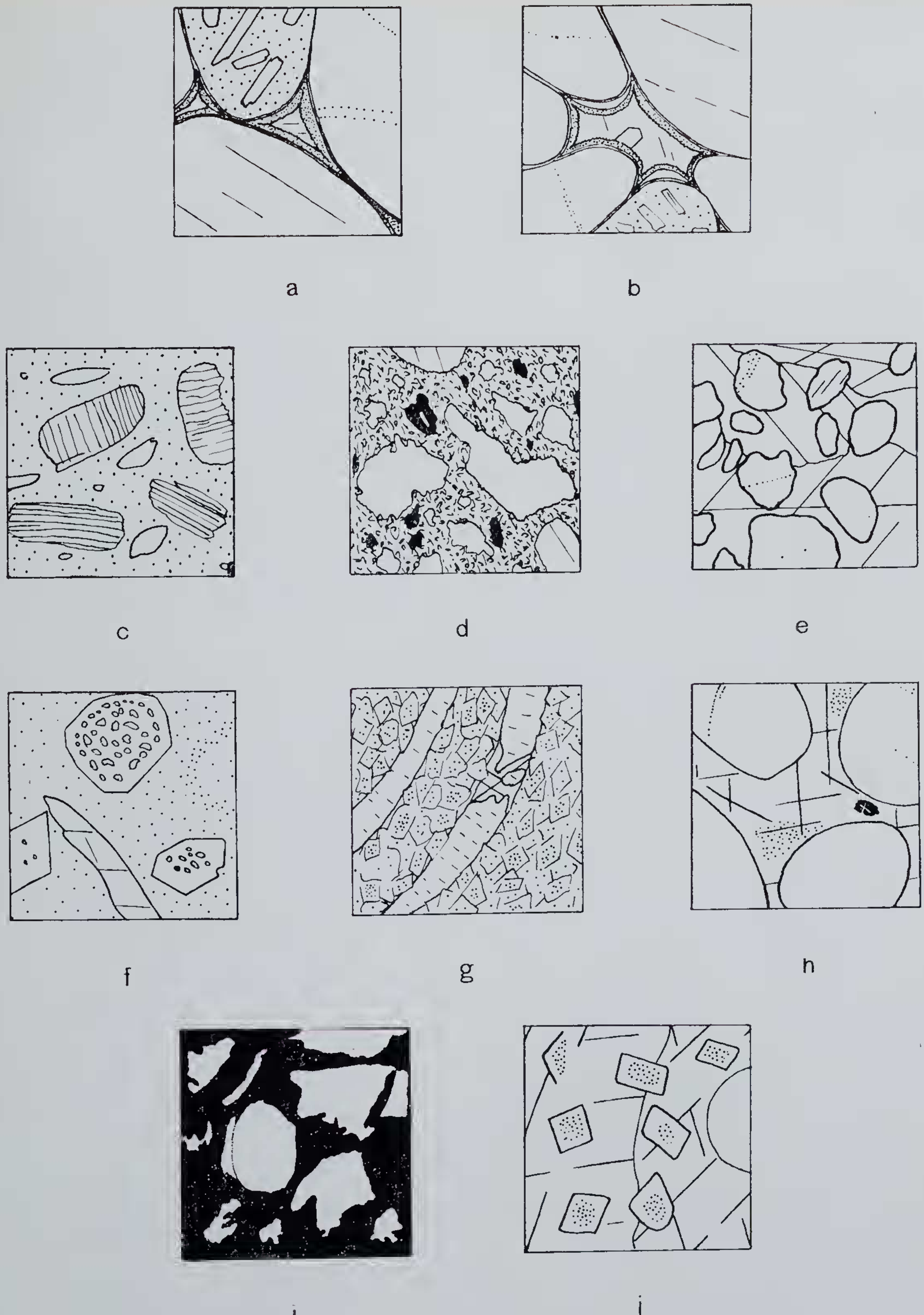

Fig. 6.

seems to accord with this origin, but there is often too much quartz to be explained by the micro-stylolites. Appeal must then be made to intrastratal solutions for at least part of the quartz. In some sandstones (better called quartzites), micro-stylolitic boundaries are confined to adjacent outgrowths, a texture probably due to imposition of great load, or faulting or folding, after diagenetic enlargement. These textures are therefore practically metamorphic. Micro-stylolites that traverse the rock, cutting across grains, are another form of pressure-solution common in limestones, but not absent from quartzites. Textures arising from pressure-solution are illustrated in Figs. 5g, 5h, 5i.

Pore-filling or Micro-drusy Textures

All micro-drusy textures are believed to form by precipitation in porous rocks, generally arenites. They differ from enlargement textures in that clastic grains do not act as nuclei on which the precipitated material grows in crystallographic continuity. This is because the clastic fragments are non-crystalline (e.g. volcanic glass, some chert), or are aggregates of minute crystals (e.g. schist, shale, fine-grained volcanic rock, ooliths, pellets, etc.) rather than single crystals, or because they are single crystals, chemically very different from the precipitated material (e.g. quartz grains and chlorite cement). The authigenic minerals may form a fringe on which the serrations are normal to grain surfaces and point into the cavity, or they may form a series of elongate crystals or fibres normal to grain surfaces and directed into the cavity. This texture is practically proof of growth in an empty pore cavity, another difference from enlargement textures, some of which are due to replacement of earlier pore-filling material.

Simple Micro-drusy Textures.—Simple textures, involving one authigenic mineral, are illustrated in Figs. 5j, 5k, 5l.

Composite Micro-drusy Textures.—Composite textures may involve several authigenic minerals, of which the outermost are the first to form, and the innermost the last. See Figs. 6a, 6b.

Reorganization Textures

Reorganization textures are like replacement textures (discussed below) except that in the former the substituted material is chemically the same as, or similar to, the original material. The reorganization is generally to a coarser crystalline form such as the change from calcilutite to sparry calcite, or from the clay-sized matrix of greywacke to aggregates of chlorite, mica, and other flaky minerals. Other changes are from finely divided clay minerals to coarse vermicular crystals, and from aragonite to calcite in fossil material, although the latter does not involve coarsening. Most reorganization textures seem to have been the result of compaction. Some of them are illustrated in Figs. 6c, 6d, 6e.

Replacement Textures

Replacement textures are very common diagenetic features of sedimentary rocks, and some are illustrated in Figs. 6f-j. They are apparently the result of almost simultaneous solution and precipitation, and in those lutites which are practically impervious to the passage of solutions, the process probably took place during compaction and the expulsion of contained fluids. As already noted, some enlargement textures can also be classified as replacement textures.

Conclusions

There is no reason to suppose that the processes leading to diagenesis are less complex than those leading to solidification of magmas, or to metamorphism. Nevertheless, they are becoming increasingly well understood, and consequently the examination of diagenetic textures should lead to petrologic inference as reliable and useful as that resulting from examination of igneous and metamorphic textures. The scheme set forth above is an attempt to group diagenetic textures so that they form a useful and simple guide to the origin of the sedimentary rocks in which they occur.

Acknowledgments

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